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J.R. WEBER, N. SANDSTROM and K.C. ARNOLD

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# GEOPHYSICAL SURVEYS ON GILMAN GLACIER, NORTHERN ILLEMERE ISLAND

**J.R. WEBER [1], N. SANDSTROM [2] and K.C. ARNOLD [3]**

## Summary

During the summers of 1957 and 1958, geophysical surveys were carried out on Gilman Glacier and the adjoining ice cap of northern Ellesmere Island. Seismic refraction and reflection measurements showed that a maximum compressional wave velocity of 3795 m/sec was reached at a depth of 50 m on Gilman Glacier. On the ice cap a maximum velocity of 3810 m/sec was reached at a depth of 100 m. Reflection profiles were calculated at various sites. On the glacier, the ice thickness varied from 380 m, 5 km from the terminus, to 760 m, 19 km from the terminus. On the ice cap, the ice thickness varied from 400 to 800 m. The mean shear stress at the bed of the glacier was found to be 0.85 bars. Calculations of the rate of discharge through a cross-section indicates that Gilman Glacier has a slightly negative regime at the present time.

During the summer of 1958, more than 200 gravity stations were established over the area of the seismic survey. The regional Bouguer anomaly was calculated from the known ice thickness at a few selected sites along the seismic profiles, and was then extrapolated for the whole area. The ice thickness was then calculated from the gravity measurements with assumed densities of 0.9 and 2.71 g. cm$^{-3}$ for ice and bedrock respectively. Agreement between bedrock profiles as determined seismically and gravimetrically was very close.

Measurements of surface movement of Gilman Glacier indicated a maximum velocity of 25 m/year.

## Résumé

Au cours des étés de 1957 et de 1958, certaines expériences géophysiques furent effectuées sur le Glacier Gilman de même que sur la calotte glaciaire attenante, sis dans le secteur septentrional de l'île Ellesmere. Les mesures de réfraction et de réflexion séismiques prises à l'intérieur du glacier Gilman établirent la vélocité maximum de l'onde de compression à 3795 m/sec à une profondeur de 50 m, tandis qu'à 100 m sous la calotte la vélocité maximum était de 3810 m/sec. Plusieurs profils de réflexion furent dressés en divers points de la région. Sur le glacier, l'épaisseur de la glace était de 380 m à 5 km du terminus, et de 760 m à 19 km du même endroit. Sur la calotte glaciaire, on a noté des épaisseurs allant de 400 à 800 m. La force moyenne de cisaillement du lit du glacier a été évaluée à 0.85 bars. De plus les calculs effectués transversalement au glacier Gilman, touchant son rythme d'ablation ont démontré que son régime actuel était quelque peu négatif.

Durant l'été de 1958, plus de 200 stations gravimétriques étaient en opération dans la région affectée aux expériences séismiques. L'anomalie régionale de Bouguer a été calculée à partir de certaines données connues sur l'épaisseur de la nappe de glace le long des divers profils séismiques, et extrapolée par la suite pour toute la région. On procéda, de plus, au calcul de l'épaisseur de la glace d'après certaines mesures gravimétriques, en assumant au préalable à la glace et à la roche des densités de 0,9 et de 2.71 g.cm$^{-3}$ respectivement. La confrontation des profils du socle rocheux établis d'après les méthodes séismiques et gravimétriques démontrèrent une très grande similarité.

Les mensurations sur le mouvement superficiel du glacier Gilman ont évalué sa vélocité maximum annuelle à 25 m.

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[1] Gravity Division, Dominion Observatory, Department of Mines and Technical Surveys, Ottawa; formerly of the Department of Physics, University of Alberta.

(2) Scripps Institution of Oceanography, University of California, La Jolla; formerly of the Geophysical Laboratory, Department of Physics, University of Toronto.

[3] Geographical Branch, Department of Mines and Technical Surveys, Ottawa; formerly employed by the Department of Physics, University of Toronto, and by the Department of Geography, McGill University.

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## 1. Introduction

The surveys described in this paper were carried out on the Defence Research Board's Operation "Hazen", the Canadian IGY expedition to the Lake Hazen area of northern Ellesmere Island, 1957-58; they covered Gilman Glacier and the ice cap in the vicinity of Mount Oxford (Fig. 1). In 1957, six reflection profiles were obtained

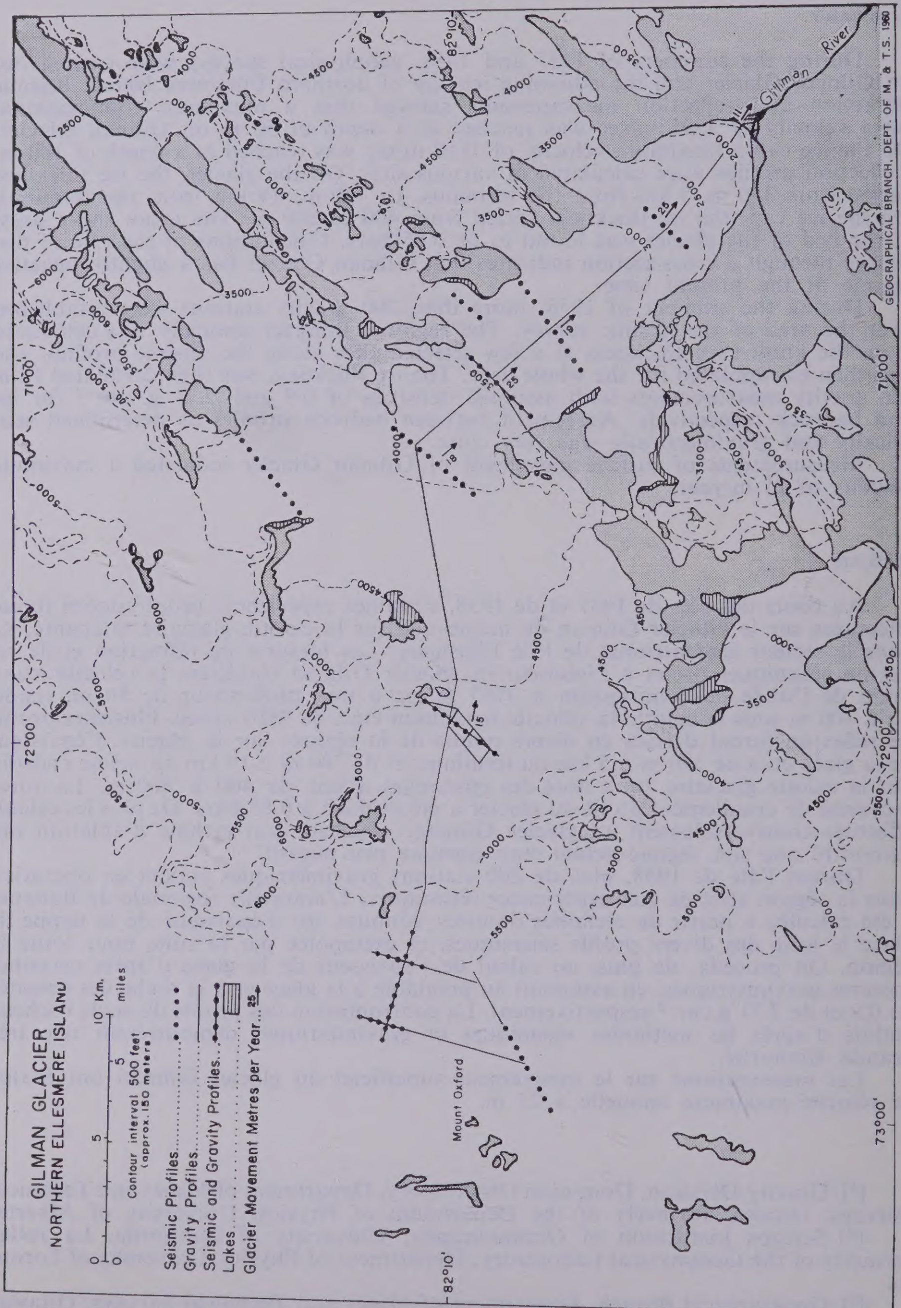
GILMAN GLACIER
NORTHERN ELLESMERE ISLAND
0 5 10 Km.
0 5 miles
Contour interval 500 feet (approx. 150 meters)
Seismic Profiles
Gravity Profiles
Seismic and Gravity Profiles
Lakes
Glacier Movement Metres per Year 25
Mount Oxford
Gilman River
82°10'
71°00'
72°00'
73°00'
GEOGRAPHICAL BRANCH, DEPT. OF M. & T.S. 1960

Fig. 1 — Map of Gilman Glacier, northern Ellesmere Island to illustrate geophysical surveys. (Prepared by the Geographical Branch, Department of Mines and Technical Surveys, Ottawa. Form lines from surveys by K.C. Arnold on Operation «Hazen»).

across Gilman Glacier and the névé at its head (Figs. 2 and 3); in 1958, a refraction profile was made on Gilman Glacier, and four refraction and reflection profiles on the ice cap. A gravity survey was undertaken over the same area in 1958, in order to determine how far depth measurements by the gravity method agreed with seismic measurements, and to supplement the data obtained from the seismic survey (Fig. 1).

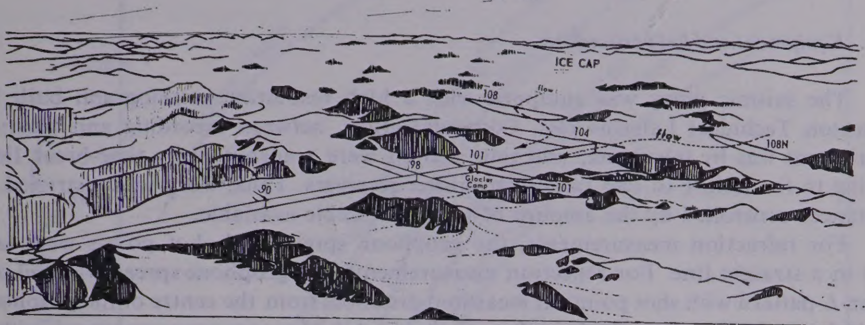

Fig. 2 — Block diagram of Gilman Glacier from the east. (From a drawing by G. Falconer).

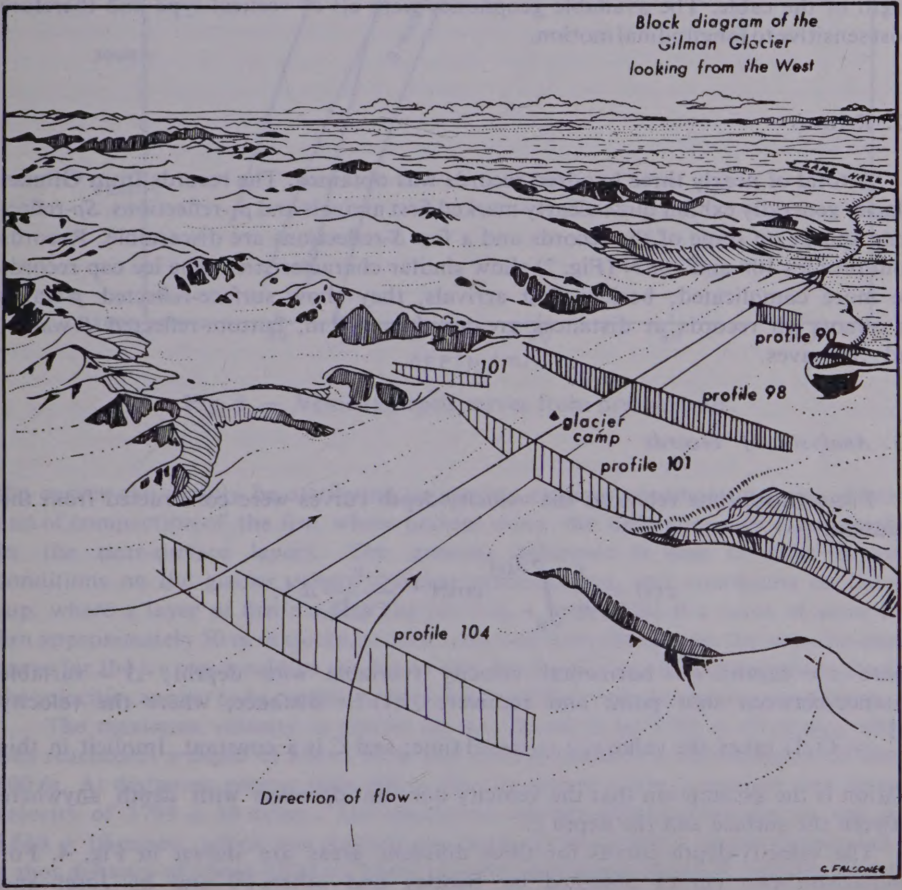

Fig. 3 — Block diagram of Gilman Glacier from the west. (From a drawing by G. Falconer).

A local network of more than 200 gravity stations was established. A triangulation and levelling survey was undertaken to establish the positions and elevations of seismic and gravity stations, and periodic observations were made on a network of stakes planted in the glacier to measure surface movement (Fig. 1).

## 2. SEISMIC SURVEY

### 2.1. *Equipment and field procedure*

The seismic party was equipped with a high resolution seismograph built by Houston Technical Laboratories. Communication between recording and blasting equipment was by telephone, and time signals were transmitted by time-break line, owing to the failure of two radio transmitter-receivers. Thus, work was restricted to distances controlled by the amount of telephone cable available.

For refraction measurements, the geophone spread and shot points were laid out in a straight line. For reflection measurements, the geophone spread was laid out in an *L*-pattern with shot points at measured distances from the centre of the geophone spread and in line with the branches of the *L*. All distances were measured with a 100-metre steel tape and all vertical angles with a Wild *T*2 theodolite. Thirty-six geophones were connected to the spread in groups of three spaced equally along the length of the cable. The available geophones were all of vertical type and therefore most sensitive to longitudinal motion.

### 2.2. *Records*

A total of nearly three hundred records was obtained. The records from Gilman Glacier generally exhibit quite clearly marked first arrivals and $p_1$-reflections. *Sp*-reflections appear on some of the records and a few *S*-reflections are discernible. Records from profiles 108 and 108 N (Fig. 2) show similar characteristics. The ice cap records are more complicated; besides first arrivals, they show surface-reflected *p*-waves on almost all records at distances greater than 400 m, bottom-reflected *P*-waves, and *Sp*-waves.

### 2.3. *Analysis of records*

From travel-time relations the velocity-depth curves were constructed from the equation:

$$z(v) = \frac{1}{\pi} \int_0^{\Delta(v)} \cosh^{-1} \frac{v}{C(\Delta)} \, d\Delta ,$$

where $z$ = depth; $v$ = horizontal velocity (variable with depth); $\Delta$ = variable distance between shot point and receiver; $\Delta(v)$ = distance, where the velocity $\frac{d\Delta}{dt} = C(\Delta)$ takes the value $v$; $t$ = travel time; and $C$ is a constant. Implicit in this relation is the assumption that the velocity does not decrease with depth anywhere between the surface and the depth $z$.

The velocity-depth curves for three different areas are shown in Fig. 4. For comparison two curves obtained by Bentley and others (1) and by Joset and Holtzscherer (2) on the Greenland ice cap are also shown. The differences between

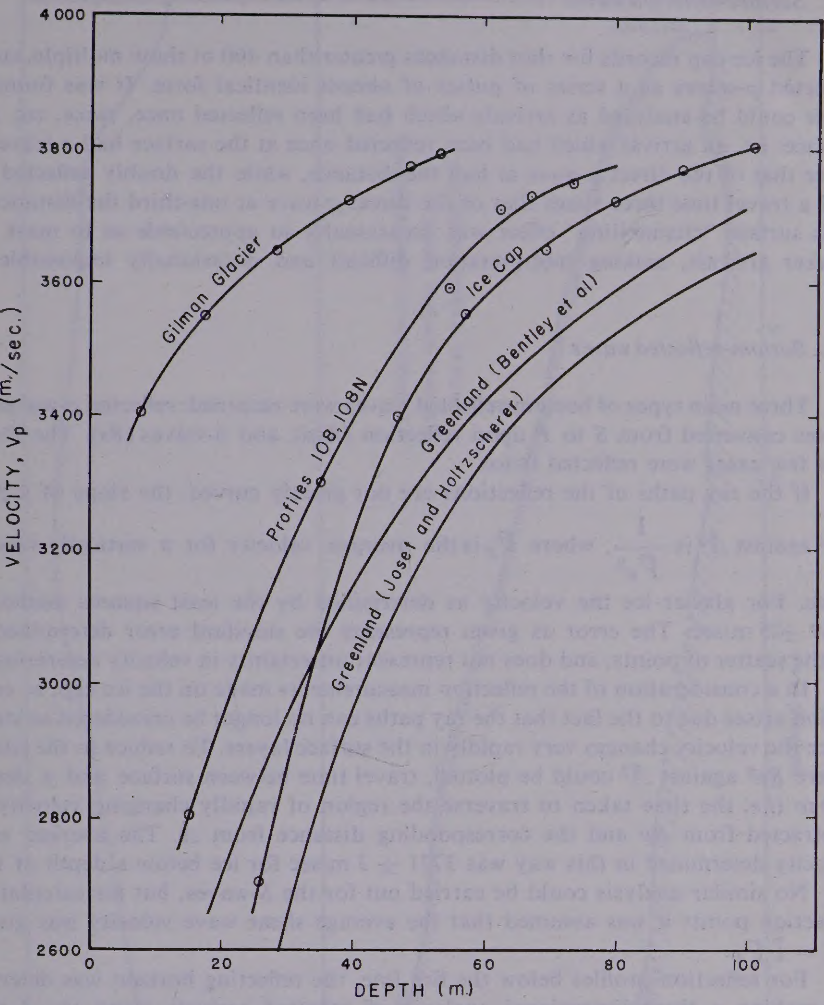

Fig. 4 — Velocity-depth curves from first arrivals.

the curves can mainly be attributed to varying rates of accumulation and ablation and of compaction of the firn where present since the difference are most marked in the near-surface layers. The greatest difference is seen to exist between conditions on the glacier, where ablation predominates, and conditions on the ice cap, where a layer of firn overlies the ice. Fig. 4 shows that if a layer of snow and firn approximately 50 m in thickness were removed from the ice cap the velocity-depth curve for the ice cap would be similar to that for the glacier. At depths below the firn the velocities appear to be mainly governed by such factors as temperature and pressure.

The maximum velocity in glacier ice was found to be 3 795 ± 10 m/sec., which was reached at a depth of about 50 m and corresponded to a shot distance of about 600 m. At distances greater than 600 m the travel-time curve is straight and gives a velocity of 3 795 ± 10 m/sec.. The maximum velocity obtained on the ice cap was 3 810 ± 10 m/sec., which was reached at a depth of about 100 m and corresponded to a shot distance of about 1 000 m. There is some uncertainty about the depths and distances quoted owing to the very slight curvature of the travel-time curve.

### 2.4. *Surface-reflected waves*

The ice cap records for shot distances greater than 400 m show multiple, surface-reflected *p*-waves as a series of pulses of almost identical form. It was found that these could be analyzed as arrivals which had been reflected once, twice, etc. at the surface: i.e. an arrival which had been reflected once at the surface had a travel time twice that of the direct *p*-wave at half the distance, while the doubly reflected wave had a travel time three times that of the direct *p*-wave at one-third the distance, etc. This surface "channelling" effect was occasionally so appreciable as to mask other weaker arrivals, making interpretation difficult and occasionally impossible (3).

### 2.5. *Bottom-reflected waves*

Three main types of bottom-reflected waves were recorded: reflected *p*-waves (*Rp*), waves converted from *S* to *P* upon reflection (*Rsp*), and *S*-waves (*Rs*). The *P*-waves in a few cases were reflected twice.

If the ray paths of the reflections are not greatly curved, the slope of a plot of $Rp^2$ against $\Delta^2$ is $\frac{1}{\bar{V}_p{}^2}$, where $\bar{V}_p$ is the average velocity for a vertically travelling wave. For glacier ice the velocity as determined by the least squares method was 3769 ± 5 m/sec. The error as given represents the standard error determined only by the scatter of points, and does not represent uncertainty in velocity determinations.

In a consideration of the reflection measurements made on the ice cap, a complication arises due to the fact that the ray paths can no longer be considered as straight, since the velocity changes very rapidly in the surface layers. To reduce to the situation where $Rp^2$ against $\Delta^2$ could be plotted, travel time between surface and a depth of 100 m (*i.e.* the time taken to traverse the region of rapidly changing velocity) was subtracted from *Rp* and the corresponding distance from $\Delta$. The average vertical velocity determined in this way was 3771 ± 3 m/sec for ice below a depth of 100 m.

No similar analysis could be carried out for the *S*-waves, but for calculation of reflection points it was assumed that the average shear wave velocity was given by $\bar{V}_s = \frac{1}{2}\bar{V}_p$.

For reflection profiles below the firn line, the reflecting horizon was determined by making a three-dimensional analysis of reflected arrivals along the *L*-shaped spread (4). The various steps in computation were programmed and performed by "Ferut", the electronic computer at the University of Toronto Computation Centre. This procedure, however, was not applicable in cases where the ray paths were considerably curved, and the data therefore insensitive to directional analysis. Thus, for the ice cap data the assumption was made that reflection occurred in a vertical plane midway between shotpoint and detector. This assumption becomes questionable when the bottom topography is undulating, and in this case corrections should be made after a first approximation for the shape of the bottom has been obtained. On Gilman Glacier a first approximation gave reasonably good indications of the topographic features beneath the ice.

The longitudinal profile of Gilman Glacier is shown in Fig. 5, constructed from information obtained by surface surveying (5) and by seismic and gravitational depth determinations. The bottom of the glacier appears rather uniform, with a gentle down-glacier slope averaging about 0.65 per cent for a distance of about 15 km, compared with a slope of 2.5 per cent at the glacier surface. The transverse sections are regular in shape and the normal cross-sections have a wide and flattened U-shape. One of the transverse sections is shown in Fig. 6. Similar profiles have been reported from one of the outlet glaciers of the Penny Ice Cap in Baffin Island (6).

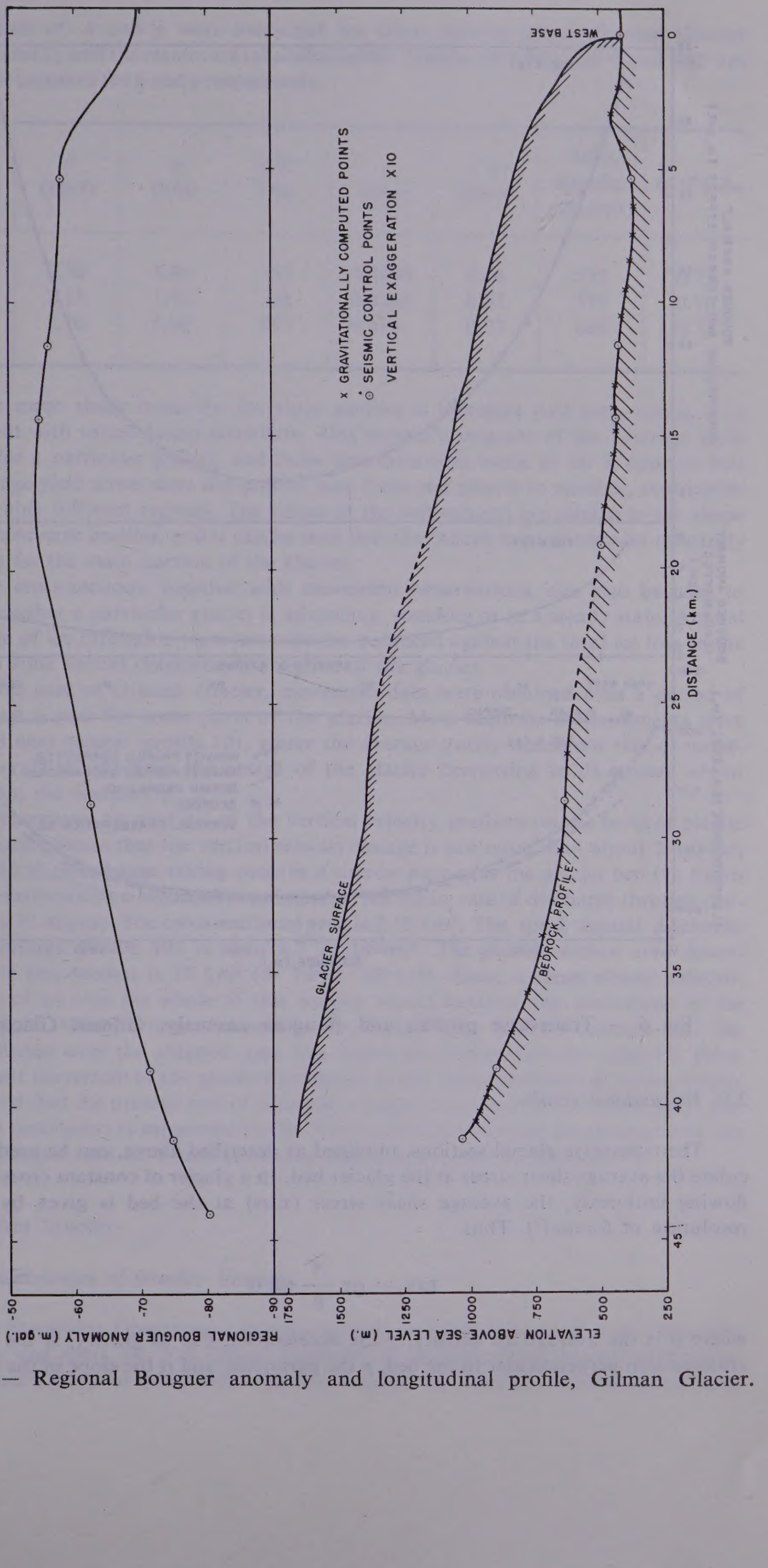

Fig. 5 — Regional Bouguer anomaly and longitudinal profile, Gilman Glacier.

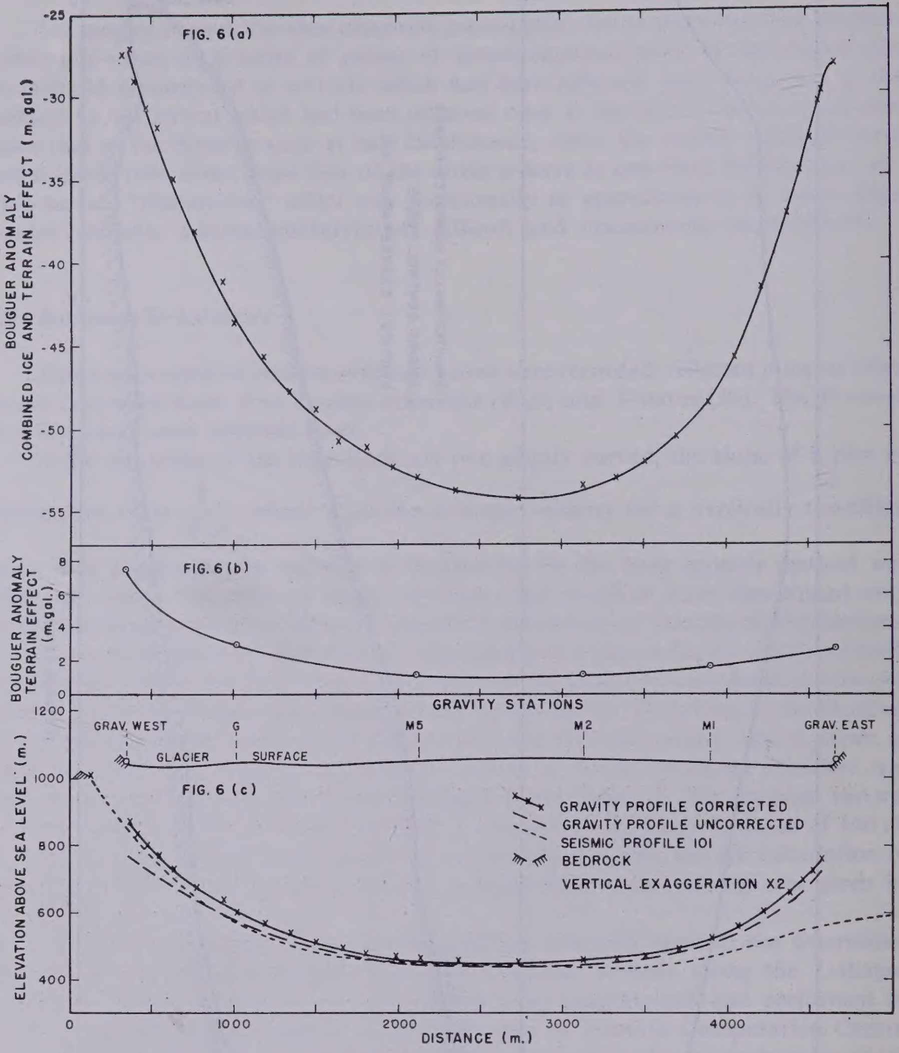

Fig. 6 — Transverse profiles and Bouguer anomaly, Gilman Glacier.

### 2.6. *Discussion of results*

The transverse glacial sections, obtained as described above, can be used to calculate the average shear stress at the glacier bed. In a glacier of constant cross-section flowing uniformly, the average shear stress $(\tau av)$ at the bed is given by simple resolution of forces (7). Thus

$$\tau av = \varrho g \frac{A}{p} \sin \alpha$$

where $\varrho$ is the average ice density, $g$ the acceleration due to gravity, $A$ the area of cross-section perpendicular to the bed, $p$ the perimeter, and $\alpha$ the slope of the surface.

Values of $A$ and $p$ were measured for three profiles across Gilman Glacier (Figs. 2 and 3), and the results are tabulated below. Values of 0.90 g $cm^{-3}$ and 985 cm $sec.^{-2}$ are assumed for $\rho$ and $g$ respectively.

| Profile | $A$ ($km^2$) | $p$ (km) | $A/P$ (m) | $\sin \alpha$ | $\tau$ av (bars) | Max. depth, $Z_m$ (m) | $(A/P)/Z_m$ |
|---|---|---|---|---|---|---|---|
| 98 | 2.30 | 5.85 | 393 | 0.0240 | 0.84 | 535 | 0.73 |
| 101 | 2.15 | 5.40 | 398 | 0.0265 | 0.93 | 575 | 0.70 |
| 104 | 2.70 | 5.60 | 481 | 0.018 | 0.77 | 660 | 0.73 |

The mean shear stress for the three profiles is therefore 0.85 bars, which is in agreement with values found elsewhere. This term is a measure of the "average yield stress" for a particular glacier, and from determinations made so far it appears that the average yield stress does not greatly vary from one glacier to another, even under considerably different regimes. The values in the last column are related to the shape of the transverse profiles, and it can be seen that the "shape factor" remains relatively constant for the main portion of the glacier.

The cross-sections, together with movement observations, can also be used to decide whether a particular glacier is advancing, receding or in a steady state. Annual discharge of ice through a transverse section balanced against the total ice loss in the ablation zone should determine the regime of the glacier.

In the case of Gilman Glacier, movement data were obtained over a period of more than a year for some parts of the glacier. Most accurate measurements were obtained near seismic profile 101, where the average yearly maximum rate of movement was 25 m/year near the centre of the glacier decreasing to 21 m/year about 1 km from the margin ($^5$).

Furthermore, an analysis of the vertical velocity gradient on the basis of plastic flow theories shows that the vertical velocity change is not more than about 2 m/year, with most of this change taking place in a narrow zone near the glacier bed ($^4$). From these considerations a reasonable estimate of the mean rate of discharge through profile 101 is 22 m/year. The cross-sectional area is 2.15 $km^2$. The total annual discharge of ice through profile 101 is then $4.7 \times 10^7$ $m^3$. The glacier surface area downstream of this section is 70 $km^2$ (= $7.0 \times 10^7$ $m^3$). Thus, a mean annual ablation of 67 cm of ice over the whole of this surface would balance the movement of ice downstream. Ablation measurements in the budget year 1957-58 indicate that the mean ablation over the ablation area was somewhat higher than this value ($^8$). Since the present movement of the glacier represents certain past conditions of stress, it may be inferred that the present rate of ablation is higher now than in the recent past; this tentative conclusion is supported by the observation that summer temperatures in the North Atlantic area have been generally higher in the last few decades ($^9$).

## 3. GRAVITY SURVEY

### 3.1. *Establishment of Gravity Stations*

The Dominion Observatory gravity network includes stations at Fort Churchill, Resolute, and Alert. Dr. F.S. Grant, a member of the Operation "Hazen" IGY team, established a series of new absolute gravity stations by tying his network into that of the Dominion Observatory. His stations included one at the base camp on Lake Hazen, one at the Gilman Glacier camp, and one on the east shore of Clements Markham Inlet, which were also stations of the local network. The local stations were chosen at sites where the bedrock elevations had been determined from previous seismic soundings ($^3$,$^4$). The gravity values were thus all established in absolute terms, although relative changes in gravity, due to the mass deficiency of the ice would have been sufficient for ice thickness computations alone. The gravimeter used was a standard Worden model. Reference was made to previous work along the same lines ($^{10}$,$^{11}$,$^{12}$,$^{13}$).

### 3.2. *Reduction of Gravity Values*

The ice thickness was computed directly on the assumption that the rock and ice under the gravity station form a slap of infinite extent. It has been shown elsewhere ($^{14}$) that the thickness ($t$) can be computed from the equation,

$$t = 1.321\,(C_B - 0.1952\,\mathrm{h}),$$

where $C_B$ = Bouguer correction for free air, rock and ice in milligals
$= \gamma - g_o - A - C_T$,
$h$ = surface elevation of the ice in metres,
$\gamma$ = theoretical gravity in milligals
$= 978{,}049\,(1 + 0.0052884\,\sin^2\varnothing - 0.0000059\,\sin^2 2\varnothing)$ where $\varnothing$ = latitude,
$g_o$ = observed gravity in milligals,
$A$ = regional Bouguer anomaly in milligals,
$C_T$ = terrain correction in milligals.

Densities for rock and ice of 2.71 and 0.9 g cm$^{-3}$ respectively were assumed.

The regional anomaly and the terrain correction are values which, in mountainous regions without exact contour maps, might be extremely difficult to estimate. However, in the case of the Gilman Glacier, because of its width, and because of the uniformity of the surface geology, the regional anomaly can be determined at its centre from a single seismic sounding. It has been shown that, since the marginal nunataks are relatively far away, the terrain correction at the centre of the glacier becomes negligible, and one depth-sounding along the centre line every few kilometres is sufficient to plot the regional anomaly curve as it changes relative to the mountain ranges (Fig. 5). The value of the regional anomaly at the location of interest can then be taken from the curve.

### 3.3. *Comparison Between Seismic and Gravity Profiles*

The ice thickness was computed for all the gravity stations and the bedrock elevations were plotted. Comparison between bedrock profiles, obtained by the seismic method, and those computed from the gravity values showed excellent agreement over the central part of the glacier. Even for profiles across the glacier the agreement was good, except at stations near the glacier edge where the discrepancy, due to the uncorrected terrain effects, became large (Fig. 6).

Some of the seismic profiles shot on the ice cap, where firn and ice was some 800 m deep ($^3$), gave very poor reflections. The reduction of the gravity observations, taken over the same profiles, revealed a very undulatory bedrock floor, which was not suspected under the level snow surface, but which was apparently the reason for the poor reflections.

### 3.4. *Conclusions*

Gravity measurements alone, without seismic depth control, can give a good picture of the shape but no accurate elevations of the bedrock. When supplemented by a few seismic soundings, the gravity method provides an excellent yet simple means of determining the shape and elevation of the bedrock beneath glaciers and ice caps. But, when no contour maps are available and terrain correction is impossible, the easy application of the method to valley glaciers is limited to geologically uniform regions with relatively wide and level glaciers. Although these conditions rarely occur in the glacierized regions of the temperate zones, they are common in the polar regions. Without elaborate terrain corrections, the gravity method fails to give the exact shape of the walls near the edge of a valley glacier.

## 4. SURVEY CONTROL

Geophysical investigations in areas not yet covered by accurate topographic maps need an adequate programme of survey control. A finer control is required for gravity than for seismic work.

For the reduction of gravity observations in the Gilman Glacier area it was necessary to establish latitudes correct to 0.1 minutes, and elevations between adjacent points correct to 0.3 m. A topographic map showing the relief of the glacier margins was required for estimating terrain corrections.

There was only one previously located point in the area. This was a "Shoran" station on Johns Island, the most northerly station of the Canadian "Shoran" network, some 45 km south-west of Gilman Glacier. The present survey was tied to this station.

The method of extended base, subtense bar traversing seemed best suited for providing survey control for geophysical profiles over wide areas of the ice cap. This method had been used previously on the inland ice of Greenland by Expeditions Polaires Françaises ($^{15}$), and by the British North Greenland Expedition ($^{16}$). It gives vertical and horizontal control of sufficient accuracy for geophysical work. The rapidity of the method and the lightness of the equipment made it well suited to a small exploratory party. The total weight of surveying equipment was 40 kg, of which no single item weighed more than 8 kg. During 1957 and 1958, a 160-km traverse was made from Chandler Fiord, 28 km south of Lake Hazen, to Clements Markham Inlet, on the Arctic Ocean 60 km north-east of Gilman Glacier; the traverse between the two sea levels closed in height to 1.2 m. Care that reciprocal vertical angles were observed simultaneously contributed to this result.

The survey programme included glacier movement observations, which were based on a local triangulation net extending up the sides of Gilman Glacier. The rate of surface movement is shown in Fig. 1. From the triangulation stations, prominent peaks were fixed, and this control formed the basis for a topographic map. The highest mountain in the area is situated about 20 km west of Mount Oxford, and its height was found to be 2500 ± 100 m. It is also believed to be the highest peak in the Queen Elizabeth Islands.

The triangulation net checked a part of the subtence bar traverse that ran from the terminus of Gilman Glacier to a small nunatak 13 km east of Mount Oxford. In a distance of 38 km the horizontal misclosure was 1 part in 3,600, and the vertical misclosure zero. This suggests that the desired accuracy was attained.

## 5. ACKNOWLEDGEMENTS

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